Chapter 9

Long-Term Eustasy and Epeirogeny

Contents

9.1 Mantle Processes and Dynamic Topography

The major cause of change in the Earth's crust is the radiogenic heat engine, which drives mantle convection and generates the geomagnetic field. Convection distributes heat and drives plate tectonics. Oceanic and continental crust are subject to heating close to sea-floor spreading centres where new, hot, oceanic crust is generated (causing "ridge push"), and to cooling overlying subduction zones, where cold oceanic crust descends to the mantle (generating "subduction pull"). Widespread cooling and subsidence occur over the downwelling zones where continental fragments converge. The same cooling and subsidence occur above old, cold, downgoing oceanic slabs in subduction zones, and partial melting of the water-saturated rocks as they descend in the subduction zone is what generates arc magmatism. The differential heat distribution and consequent heating and cooling of different parts of the overlying Earth's surface results in broad regional uplifts, downwarps and tilts, because of the effects on crustal densities. These processes maintain what is called *dynamic topography* (Fig. 9.1). The effects of crustal heating are to cause uplift. This occurs along the flanks of new continental rift systems (e.g., parts of East Africa) and above mantle plumes. Thermal doming beneath supercontinents may elevate the crust by as much as 1 km over periods of 100 million years. Subsidence takes place over cooling areas of the Earth's crust, such as areas of aging oceanic and continental crust distant from spreading centres, and over regions of mantle downwelling.

The concepts of dynamic topography are now being explored with the techniques of a new field, called computational geodynamics, "in which computer models of mantle convection are used in the interpretation of contemporaneous geophysical observations like seismic tomography and the geoid as well as of time-integrated observations from isotope geochemistry" (Gurnis, 1992). Such models are capable of integrating large volumes of detailed stratigraphic data using the backstripping procedures outlined in Sect. 3.5. These developments are of great significance, and their implications for sequence stratigraphy have yet to be fully realized.

Vertical movement of the crust causes relative changes in sea level on a regional or continental scale. This is true *epeirogeny*. However, thermal changes also result in changes in the volume of the ocean basins, which lead, in turn, to eustatic changes in sea level. The same broad crust-mantle processes generate both regional and eustatic effects that may or may not be in phase. The result is a highly complex sequence of sea-level changes, and clear global eustatic signals may not always be present in the stratigraphic record.

Continental-scale vertical movements caused by the process of epeirogeny have been amply demonstrated by stratigraphers (e.g., Sloss, 1963; Bond, 1976, 1978; Burgess, 2008; see Sect. 3.4), but were only explained

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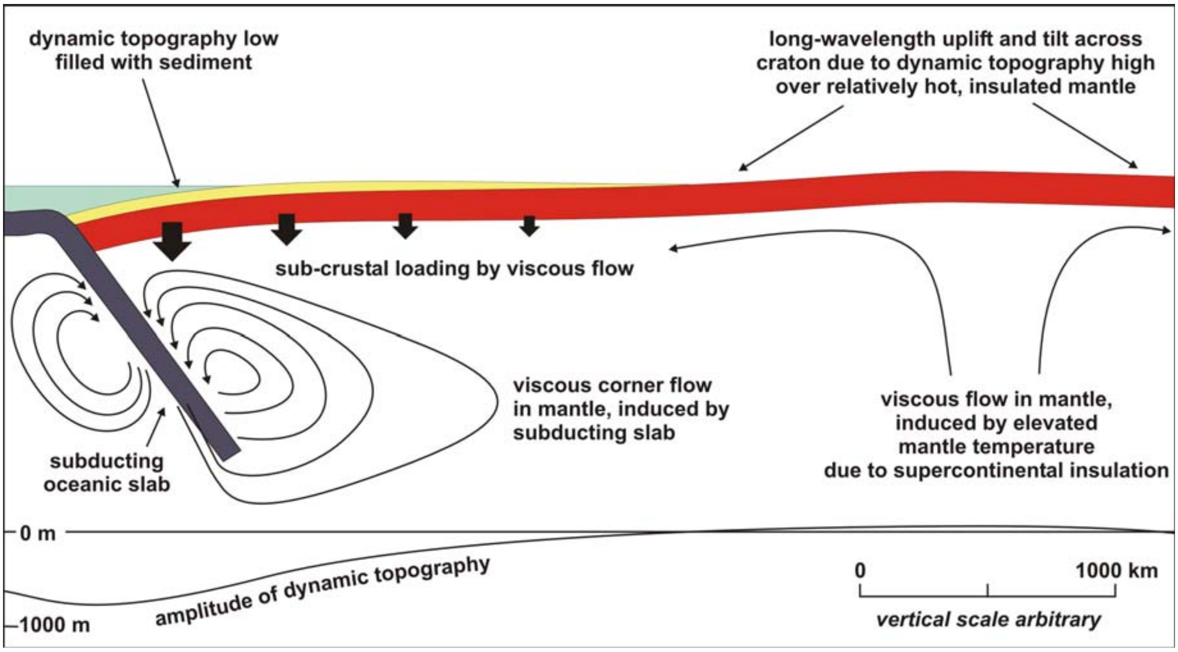


Fig. 9.1 The generation of dynamic topography and controls on sea-level change and regional vertical crustal motion by thermal effects related to shallow and deep mantle-convection. Widespread heating beneath supercontinents (at *right*) generates continental uplift and an elevated geoid. More localized depression of the geoid is caused by the subduction of cold slabs of oceanic crust (at *left*) (Burgess, 2008, Fig. 6)

recently. At a time when the global eustasy model remained popular, Sloss (1991) maintained that vertical and tilting movements driven by epeirogeny were indicated by the large-scale architecture of the 10^7-year sequences (the "Sloss sequences"). His cross-sections (e.g., Fig. 5.8) and his maps of subsidence patters of the craton beneath the United States (e.g., Figs. 5.9 and 5.10) make this point, and it is one reinforced by the discussion of the Phanerozoic evolution of the craton by Burgess (2008), who stated (p. 37):

> . . . there is a very basic observation that demonstrates that eustasy was not the only contributor to the relative sea-level changes recorded by cratonic sequences. Long-term eustatic oscillations certainly must have contributed to development of the transgressive and regressive sequence elements, but long-wavelength post-depositional tilting of the cratonic strata and angular sequence-bounding unconformities, both ubiquitous features of North American cratonic strata . . ., obviously require a tectonic mechanism, and cannot be explained by eustatic change alone. Applying such simple reasoning to North American cratonic stratal patterns often provides a reasonable indication of the degree of tectonic and eustatic influence on relative sea-level.

Only in the last two decades has this research finally provided a theoretical basis for the process of epeirogeny, a process that has commonly been invoked to explain the broad vertical movements of the crust indicated by stratigraphic studies, but which has lacked a basis in the modern theories of plate tectonics.

The processes referred to here are relatively long-term in their effect, and are capable of explaining much of the cyclicity that has been recorded on time scales of tens to hundreds of millions of years. They are the subject of this chapter.

9.2 Supercontinent Cycles

Early work by Wilson, Sutton, Bott, Condie, Windley, and others suggested that the Phanerozoic history of the earth (the broad, long-term stratigraphic patterns outlined in Sect. 5.2) can be related to the assembly and breakup of the Pangea supercontinent. Most recent workers have adopted this long-term plate-tectonic cycle as the basis for hypotheses of the earth's dynamics (Anderson, 1982, 1984, 1994; Worsley et al., 1984, 1986; Gurnis, 1988; Veevers, 1990; Dewey and Pitman, 1998).

The formation of a supercontinent creates a thermal blanket that inhibits convective release of radiogenic heat from the mantle (Fig. 9.1). Changes in the rotation of the Earth's core and in the convective patterns in the mantle may be either the cause or the consequence of these surface events, which also appear to be linked to changes in the earth's magnetic field (Anderson, 1984; Maxwell, 1984). Development of the thermal blanket may be the cause of the eventual breakup of the supercontinent, following establishment of a new pattern of mantle convection. The formation of the thermal blanket beneath a supercontinent leads to heating and regional epeirogenic uplift on a continental scale. Gurnis (1988, 1990, 1992) demonstrated that the uplift rate would be 5–10 m/million years, and could persist for 100 million years, resulting in an uplift of 500 m to 1 km. (This is the first of many processes of sea-level change discussed in this book for which quantitative estimates of rate, duration and magnitude are available. The processes are listed in Table 8.2, which is referred to throughout the remainder of this book). A summary of the process of supercontinent assembly and fragmentation is provided in Sect. 5.2 (see also Fig. 9.2).

It has been argued that dynamic mantle uplift is extremely long-lived. It generates a positive geoid anomaly that survives for 10^8 years. Crough and Jurdy (1980) demonstrated the existence of a large positive anomaly beneath Africa. Veevers (1990) showed the position of this anomaly beneath a reconstruction of the Pangea supercontinent. The correspondence is remarkably close, and confirms that Africa was at the centre of Pangea. As noted in Sect. 3.4, Bond (1976, 1978) has demonstrated that Africa has undergone anomalous uplift since the early Tertiary. This is too late to have been caused directly by the heating effect, which would have taken place in the late Paleozoic or early Mesozoic following continental assembly. However, it is possible that the uplift relates to intraplate compressive stress generated by the opening of oceans virtually all around the continent. Dispersing continental fragments tend to migrate toward geoid lows, where mantle temperatures are lower, and where relative sea-levels will rise, leading to extensive platform flooding (Gurnis, 1988, 1990, 1992).

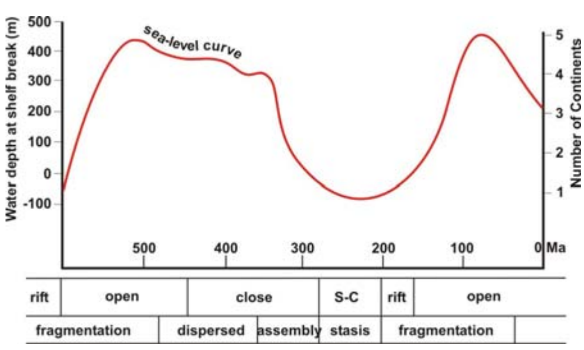


Fig. 9.2 The supercontinent cycle, showing its relationship to long-term changes in sea level. The "stasis" phase corresponds to the main period of supercontinent assembly (S-C), which was Pangea in the diagram above (simplified from Fig. 5.2)

The total length of rifting continental margins and of seafloor spreading centres increases during the breakup of a supercontinent and may be accompanied by increased rates of oceanic crust generation, and active subduction, plutonism, and arc volcanism on the outer, convergent plate margins of the dispersing fragments, as suggested by the westward drift of the Americas since the Triassic, and the consequent subduction of tens of thousands of kilometres of paleo-Pacific (Panthalassa) oceanic crust (Engebretson et al., 1985). Spreading rates are episodic, reflecting the structure and behaviour of the mantle convection cells that drive them (Gurnis, 1988). Major eustatic transgressions occur because of the displacement of ocean waters by thermally elevated young oceanic crust and active spreading centers in the new Atlantic-type oceans. These factors were considered by Pitman (1978), Kominz (1984) and Harrison (1990) in the development of a model for cycles of eustatic sea-level changes over time periods of tens to hundreds of millions of years (see next section).

Heller and Angevine (1985) argued that during the first 50–100 million years after initiation of breakup of a supercontinent, the global average age of oceanic crust decreases because of the active development of Atlantic-type oceans. This will lead to a rise in sea-level, without any change in global average spreading rate. Dockal and Worsley (1991) developed a simple isostatic model to quantify the effects of changing age of the earth's oceanic crust, simplifying the earth to a two-ocean system, with an opening Atlantic-type ocean replacing a Pacific-type ocean undergoing consumption. They demonstrated that this effect alone can account for a change in sea level of ~100 m over about 120 million years (Table 8.2).

It has been suggested that the average rate of spreading slows at times of continental assembly, at a time of ridge reordering following major continental collision and suturing events. Collision results in crustal shortening and thickening, which has the effect of increasing the ocean-basin volume. Therefore, at the end of a supercontinent assembly cycle, large areas of old, and therefore cool, and subsided oceanic crust will underlie the world's oceans (Worsley et al., 1984, 1986). All these effects lead to enlargement of the world's ocean basins. Times of low sea-level might therefore be expected to correlate with, or follow, major suturing episodes (Valentine and Moores, 1970, 1972; Larson and Pitman, 1972; Vail et al., 1977; Schwan, 1980; Heller and Angevine, 1985).

The effects of sea-level change on climates, sedimentation and biogenesis are briefly described in Sect. 5.1.

Although eustatic sea level is predicted to fall during continental assembly, Kominz and Bond (1991) documented a synchronous rise in relative sea-level in intracratonic basins and continental margins throughout North America during the middle Paleozoic (Late Devonian-mid Mississippian) (Fig. 9.3), at which time it is postulated that the late Proterozoic supercontinent was dispersing, and Pangea was assembling (Worsley et al., 1984, 1986). Kominz and Bond (1991) attributed the regional rise in sea level to synchronous enhanced subsidence, and argued that this could not have been caused directly by plate-margin processes. Many of the basins are beyond the flexural reach of the continental-margin orogenies that were underway at the time, and some of the data were derived from areas undergoing continental extension where no thermal event has been documented that could explain the timing or rate of subsidence. The synchronous nature of the subsidence (Fig. 9.3) calls for a continental-scale process, and Kominz and Bond (1991) suggested that the cause was intraplate compressive stress resulting from the movement of the North American plate over a region of mantle downwelling during supercontinent dispersion. Kominz and Bond (1991, p. 59) stated:

> . . . the converging limbs of the convection system would increase the in-plane compressive stresses at the base of the lithosphere. For a critical level of stress, all preexisting positive and negative lithospheric deflections (i.e., arches and basins) will be enhanced; arches will tend to move upward and basins will tend to subside. The convection modeling predicts that the maximum compressive stress in a downwelling region is 60 to 70 MPa (Gurnis, 1988), a range that probably is sufficient to reactivate preexisting deflections, assuming a viscoelastic lithospheric rheology.

Subsidence over downwelling mantle is an expression of "dynamic topography", a process described in Sect. 9.3.2. The hypothesis of intraplate stress has been developed by Cloetingh (1986, 1988). It may be responsible for regional changes in relative sea level over time scales of tens of thousands to tens of millions of years, and is discussed at greater length in Sect. 10.4.

In the case of the North American sea-level rise in the Middle Paleozoic, subsidence would have been enhanced by the increase in crustal density overlying a cool downwelling current (the dynamic topography effect). It is not yet clear how much this effect contributed to the overall relative sea-level rise.

As pointed out by Gurnis (1992), there is an ambiguity in attributing causes to long-term sea-level changes. Mantle convection leads to generation of dynamic topographies, which are reflected in the stratigraphic record by their continental-scale effects on relative sea levels. However, the same processes lead to changes in the global average rate of sea-floor spreading, which affect the volume of the ocean basins, and thereby generate eustatic sea-level changes. During times of supercontinent fragmentation, in areas of mantle downwelling, these two processes will be in phase, and therefore additive, which makes it difficult to separate and quantify their effects.

9.3 Cycles with Episodicities of Tens of Millions of Years

9.3.1 Eustasy

Hallam (1963) suggested that eustatic sea-level oscillations could be caused by variations in oceanic ridge volumes. Later workers (e.g. Russell, 1968; Valentine and Moores, 1970, 1972; Rona, 1973; Hays and Pitman, 1973; Pitman, 1978) applied the increasing knowledge of plate-tectonic processes to suggest that variations in seafloor spreading rates, variations in total ridge length, or both are the cause of the volume

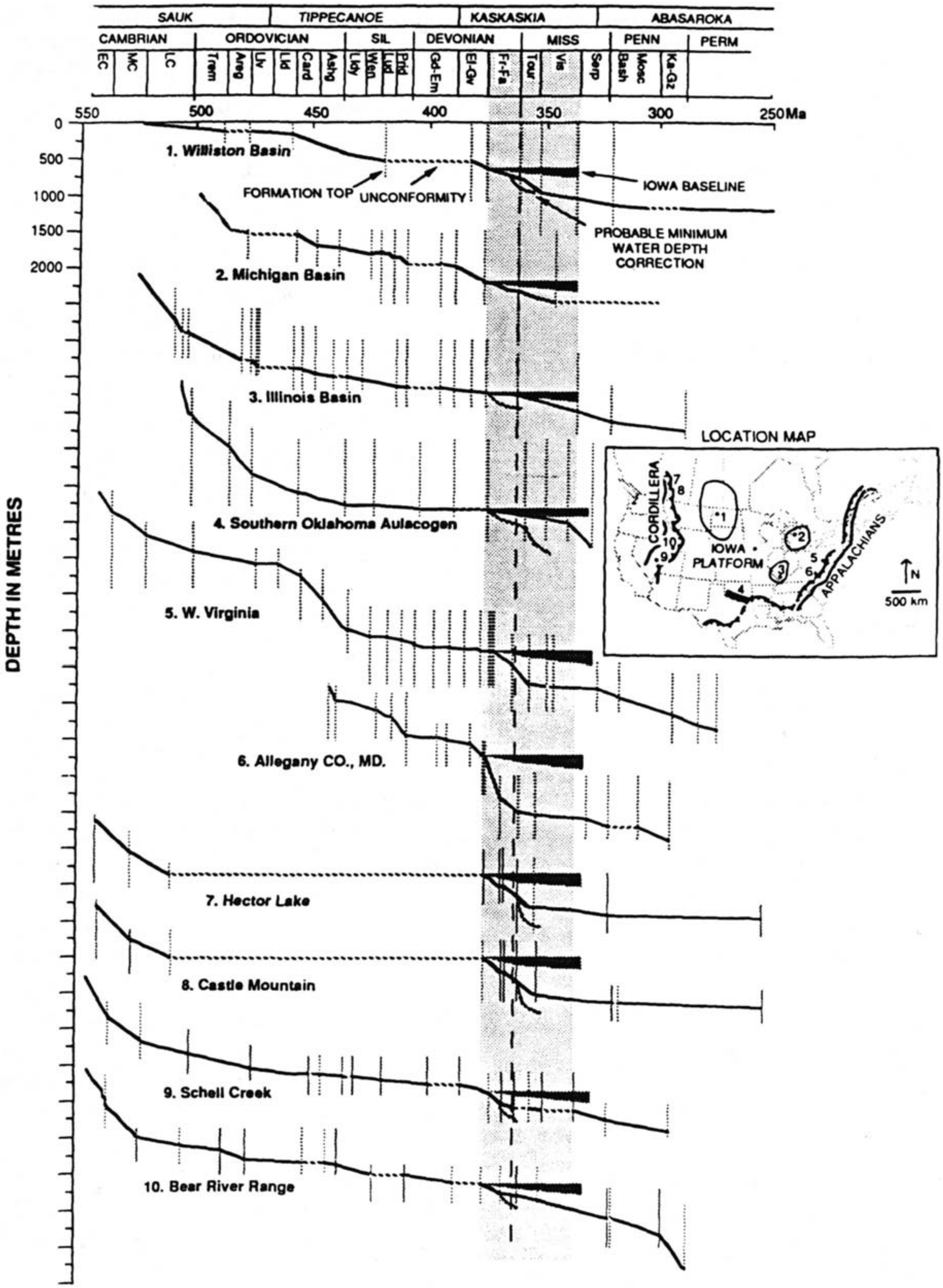


Fig. 9.3 Subsidence curves from various locations in North America, showing period of accelerated subsidence in the Late Devonian to middle Mississippian (Kominz and Bond, 1991)

changes. The average age of the oceanic crust also changes, especially during the assembly and dispersal of supercontinents, as noted in the previous section, and this also affects the volume of the ocean basins.

The oceanic lithosphere formed at a spreading center is initially hot, and cools as it moves away from the axis. Cooling is accompanied by thermal contraction and subsidence (Sclater et al., 1971). The age-versus-depth relationship is constant, regardless of spreading history and follows a time-dependent exponential cooling curve (McKenzie and Sclater, 1971), as does the overlying continental crust.

Lowstands of sea level would occur during episodes of slow spreading, during which relatively small volumes of hot oceanic lithosphere are being generated. Conversely, episodes of fast spreading would raise sea levels by increasing the ridge volumes. Using the data of Sclater et al. (1971), Pitman (1978) modeled volume changes in a hypothetical ridge, as shown in Fig. 9.4.

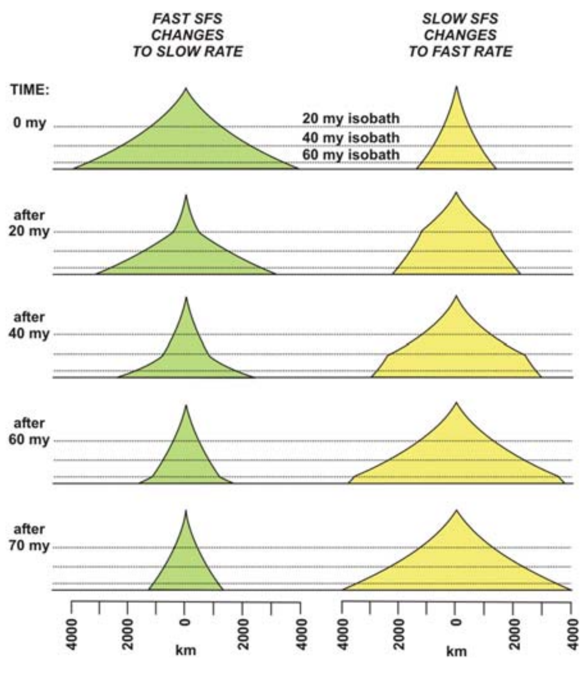


Fig. 9.4 Profiles of spreading ridges showing the effect of different spreading rates on volume at 20, 40, 60, and 70 million years after initial condition. Right: Profile of a ridge that has been spreading at 20 m/ka for 70 million years and changes to 60 m/ka at time zero. After 70 million years, the ridge has three times its starting volume. Left: Ridge that has been spreading at 60 m/ka for 70 million years and changes to 20 m/ka at time zero. After 70 million years its volume has been reduced to one third (Pitman, 1978, 1979)

The elevation of any part of a ridge can be calculated by converting age to depth, using an appropriate spreading rate. Pitman (1978) showed that a ridge spreading at 60 m/ka will have three times the volume of one spreading at 20 m/ka, provided these rates last for 70 million years (Fig. 9.4). This is the time taken for the oldest (outermost) part of the ridge to subside to average oceanic abyssal depths of 5.5 km, by which time the ridge has achieved an equilibrium profile. The total length of the world midoceanic ridge system is about 45,000 km (Hays and Pitman, 1973; Pitman, 1978), and Pitman (1978) argued that, allowing for the shape of the continental margins, measured spreading rates can account for eustatic sea-level changes up to a maximum rate of 0.01 m/ka. Later compilations (e.g., Pitman and Golovchenko, 1991; Dewey and Pitman, 1998) demonstrated average rates of 0.002–0.003 m/ka for the Jurassic to mid-Tertiary. This is fast enough to generate Sloss-type cycles, those with episodicities of tens of millions of years.

A rise or fall in sea-level is not necessarily the same thing as a transgression or regression. As the rifted margins of a continent move away from a spreading center, they subside as a result of thermal contraction, crustal attenuation, and possibly, phase changes (Sleep, 1971; Watts and Ryan, 1976). Sediments deposited on the subsiding margin cause further isostatic subsidence. The stretched and loaded margin rotates downward around a hinge located at or near the inboard limit of stretched crust. Immediately after rifting and the appearance of oceanic crust, the margins may subside at rates in the order of 0.2 m/ka, decreasing after a few million years to about 0.03–0.07 m/ka and to less than 0.03 m/ka after about 20 million years (Watts and Ryan, 1976; McKenzie, 1978; Pitman, 1978). Average rates of tectonic subsidence on these trailing margins calculated by backstripping procedures are little more than 0.01 m/ka (Pitman and Golovchenko, 1983). Shelf-edge subsidence appears to be always slightly faster than the rate of long-term eustatic sea-level change caused by changes in ridge volumes, and therefore transgressions can actually occur during periods of falling eustatic sea level, if the rate of lowering is sufficiently slow. Conversely, regressions can occur locally during periods of rising sea level if there is an adequate sediment supply. Growth of large deltas (such as that of the Mississippi) following the Holocene postglacial sea-level rise is adequate demonstration of this. However, such local regressions are not relevant to the consideration of global stratigraphic cycles. Except on continental margins underlain by unusually old, rigid crust, which subside slowly, it is unlikely that sea-level changes caused by ridge-volume changes could lower sea-level to beyond the shelf-break (Pitman and Golovchenko, 1983). These tectonic processes that control relative sea level on a regional scale are discussed further in Chap. 10.

The spreading histories of the world oceans are now reasonably well understood (including the recent synthesis of the history of the Arctic regions by Lawver et al., 2002), based on deep-sea-drilling and magnetic-reversal data. Knowledge of worldwide spreading rates enabled Hays and Pitman (1973) to calculate ridge-volume changes and a sea-level curve for the last 110 Ma. A revised version of this curve was calculated for the period 85–15 Ma by Pitman (1978), based on refinements in oceanic data. It showed a gradual drop in sea level of 350 m at an average rate of 0.005 m/ka. Sea levels rose to a maximum during a period of fast spreading between 110 and 85 Ma (Larson and Pitman, 1972), and the subsequent drop reflects slower spreading rates. By calculating the relationship between spreading rates, subsidence rates, and falling sea-level, Pitman (1978) was able to model a major global transgression during the Eocene, as actually documented from stratigraphic evidence by Hallam (1963), and a second transgression during the Miocene. An early Oligocene regression is probably related to the onset of large-scale Antarctic glaciation (Fig. 4.10). Vail et al. (1977) used Pitman's curve to calibrate their chart of relative changes of sea level, in the belief that in this way they were adjusting the curve to show true eustatic sea-level change. They suggested that positive departures from Pitman's curve (where Vail's curve shows a higher sea level than Pitman) are the result of temporary increases in subsidence rates because of sediment loading. Negative departures were attributed to rapid sea-level falls driven by glacioeustasy, a process not factored into Pitman's curve (Vail et al., 1977, p. 92). It is of interest that Vail et al. (1977) mentioned tectonic subsidence as a factor in generating onlap and relative rises in sea level, because the general, indeed overriding importance of this process (discussed in Chap. 11) was been almost completely ignored in subsequent work by the Exxon group, until very recently.

Kominz (1984) reexamined the data on which Pitman's (1978, 1979) curve was based, carrying out her own calculations and incorporating new data, and showed that, because of an incomplete data base, a considerable error must still be allowed for in the development and use of a long-term sea-level curve. She used arbitrary, estimated spreading rates for the Tethyan Ocean because, of course, this ocean has now been completely subducted, whereas during the Mesozoic and early Tertiary it was one of the world's major oceans and its sea-floor spreading history would have had a considerable effect on the eustatic curve. Other errors include inaccurate dating of the sea floor, and missing data; for example the spreading history of the Arctic Ocean was unknown at the time of her synthesis (and is still incompletely understood). Kominz (1984) concluded that the range of error is ∼120 m at 80 Ma, decreasing to ∼10 m at present. This has important implications for the accuracy of backstripping calculations used to reconstruct basin subsidence histories. More recent work by Kominz and her colleagues on the sea-level record is discussed below and in Sect. 14.6.

A compilation of spreading centres associated with the breakup of Pangea is shown in Figs. 9.5, and 9.6 shows the age of initiation of rifting along each segment of these breakup fractures. The initiation of each ridge would have had an effect on the global average spreading rates, and on the age distribution of the oceanic crust, with consequences for eustatic fluctuations over time periods of tens of millions of years. However, as discussed in Chap. 10, the same events are now also thought to have had a profound effect on regional intraplate-stress regimes, with consequences for regional warping and tilting, and the generation of relative sea-level changes over large continental areas. Rifting and continental separation is also one of the major processes involved in the development of *tectonostratigraphic sequences* (Sect. 10.2).

Estimates of the rate and magnitude of eustatic sea-level changes that can be attributed to volume changes in sea-floor spreading centres were made by Pitman and Golovchenko (1991), based on their earlier work and that of Kominz (1984), and are given in Table 8.2. Dockal and Worsley (1991) examined the effects on the age distribution of the Earth's oceanic crust during the formation and breakup of supercontinents. As noted in the previous section, a two-ocean model, in which a Pacific-type ocean (Panthalassa) is replaced by an Atlantic-type ocean, can account for tens of metres of sea-level change over a time scale of hundreds of

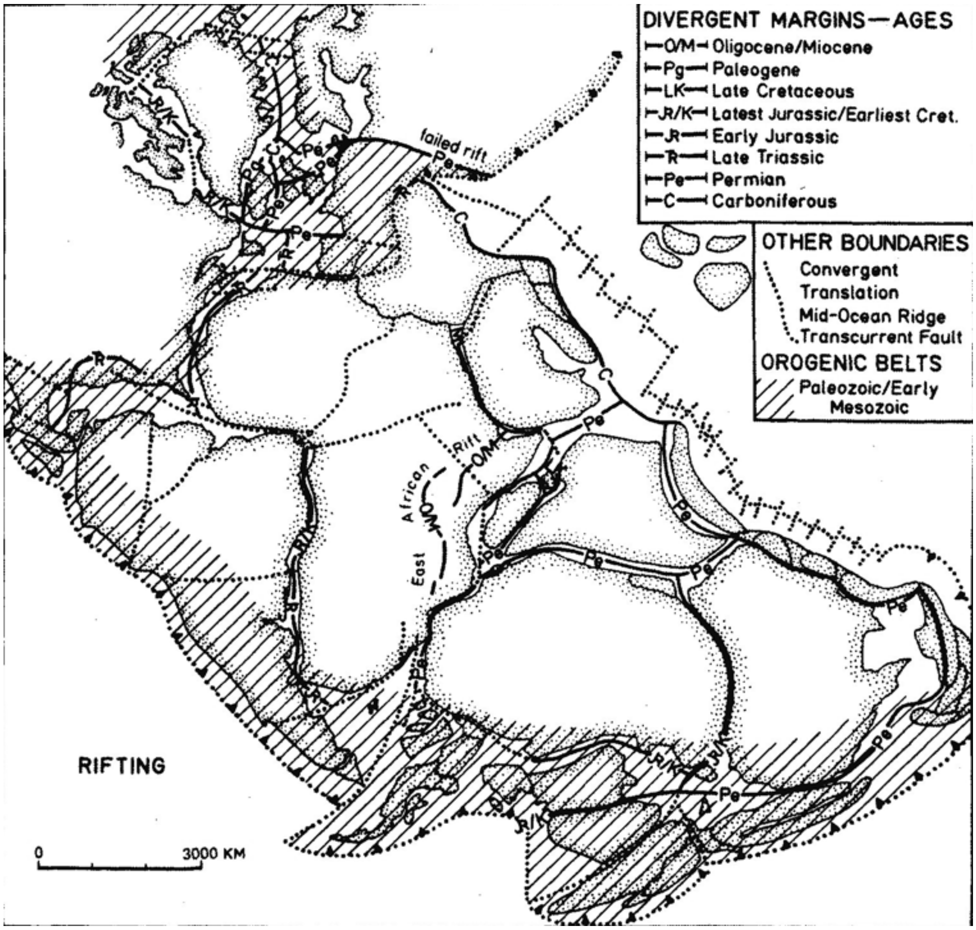


Fig. 9.5 Paleogeography of Pangea during the earliest Triassic, showing the distribution and ages of initiation of rifting on what became divergent continental margins during the Mesozoic and Cenozoic (Uchupi and Emery, 1991)

millions of years. Preliminary modeling of more complex oceanic systems, such as a two-phase opening of the Atlantic Ocean, indicated that second-order effects would also occur with an amplitude up to about 10 m. The opening of other, smaller oceans (e.g., Labrador Sea, Red Sea) would have had similar, if smaller effects. Dewey and Pitman (1998) made other calculation, including the effects on the ocean-basin volumes of the large-scale continental compression and contraction that occurs during collision and suturing, such as that of India against Asia, and the effects of the filling of large-scale depressions in the Earth's crust, such as that represented by the Mediterranean basin, which desiccated and dried out completely during the Miocene. Their estimates of the effects on eustatic sea level are included in Table 8.2.

Other processes that could possibly affect eustatic sea levels on a time scale of tens to hundreds of millions of years were reviewed by Harrison (1990), who calculated the effects on sea level by relating them to changes in the total volume of the world ridge system. Continental collision increases the thickness of the crust and decreases its area, thus increasing the area and volume of the ocean basins. Major collision and shortening events, such as that between India and Asia, therefore result in a lowering of sea level. The generation and subsequent cooling and subsidence of large oceanic volcanic extrusive masses can be shown to have a modest effect on sea level. Sediment deposited in oceans has an isostatic loading effect which amplifies the subsidence due to crustal aging, but also displaces water. Changes in global average ocean

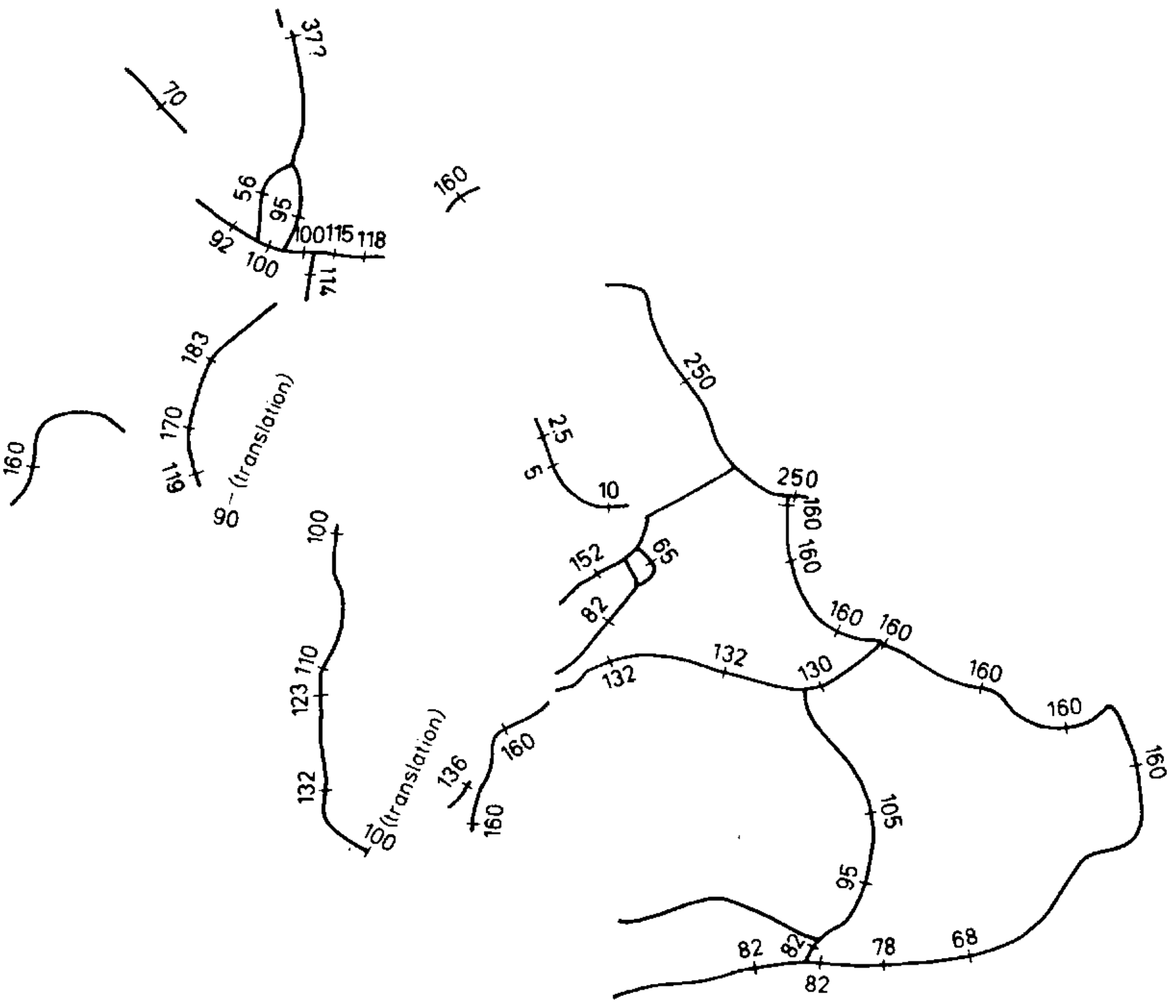


Fig. 9.6 Divergent continental margins that developed within and around Pangea, showing the age of initiation of seafloor spreading (Uchupi and Emery, 1991)

temperature change the volume of the water through thermal expansion and contraction, without changing its load. Significant temperature changes have occurred as the Earth has cycled between icehouse and greenhouse states. Fairbridge (1961) estimated that a 1° rise in ocean temperatures would lead to a 2 m rise in sea level. Harrison (1990) and Dewey and Pitman (1998) calculated this effect at 10 m.

Modern work on long-term sea-level changes has been discussed by Dewey and Pitman (1998), Müller et al. (2008) and Kominz et al. (2008). As Dewey and Pitman (1998, p. 10) noted:

> ... the rise and fall of sea level can only be viewed rationally in the context of major tectonic episodicity involving the assembly, dismemberment and dispersal of the larger continental masses in "Pangea"-type supercontinent megacycles, where tectonics and sea level are closely interrelated with much buffering, feedback loops and enhancement. This further involves a close relationship between continental distribution, plate kinematics and state of stress, climate, geomorphology, sedimentary facies and basin stratigraphy.

Müller et al. (2008) presented a comprehensive reconstruction of the global age-area and depth-area distribution of ocean floor, including remnants of subducted crust, since the Early Cretaceous (140 Ma), to compute the effects of changes in crustal production, sediment thickness, and ocean-basin depth and area on sea-level fluctuations through time. They made use of recent data regarding the age distribution of oceanic crust, the possible effect of the eruptions associated with large igneous provinces (the effect of which they estimate could raise sea level by as much as 5 m), and the effects of the subduction of oceanic crust. The latter, by cooling the overlying mantle, could cause widespread subsidence of the overlying crust. This is the process known as dynamic topography, and is discussed in the next section.

A range of modern data sets, including those of Müller et al. (2008), was compiled by Kominz et al. (2008), the results of which are shown in Figs. 9.7 and 9.8. The curves shown in Fig. 9.7 were compiled using various methods. Kominz (1984) analyzed the effect of sea-floor spreading on ridge volumes and, thus, eustasy, and indicated the large range of potential error associated with the estimates, based on the knowledge available at that time concerning ocean-floor history,. Two of her plots are provided, the range of sea level, assuming minimum spreading rates (Bio) for the Cretaceous quiet period (green band), and the mean, best estimate result (b.e.) curve. Müller et al. (2008) updated these curves by analyzing ocean volume change including the effect of ridge volume, sediment volume, large igneous province emplacement, and the changing area of the oceans. Ice is not included in their results, which are plotted assuming an ice free world (54 m above today's sea level). Also included in these figures are some older estimates. Bond (1979) studied global continental flooding history and estimated a range of possible sea level for broad periods of time. Sahagian et al. (1996) carried out a detailed analysis of Cretaceous and older sea level on the stable Russian Platform. Watts and Thorne (1984) combined backstripping and forward modelling of the Atlantic coastal stratigraphy to derive their sea level estimates. The Haq et al. (1987) long-term and short-term curves are included, but lie somewhat to the right of the other curves because their estimates for sea-level fluctuations were substantially larger that those of most other workers, despite being calibrated against the best-estimate values of Kominz (1984)—note the coincidence of these two curves at 230 m at

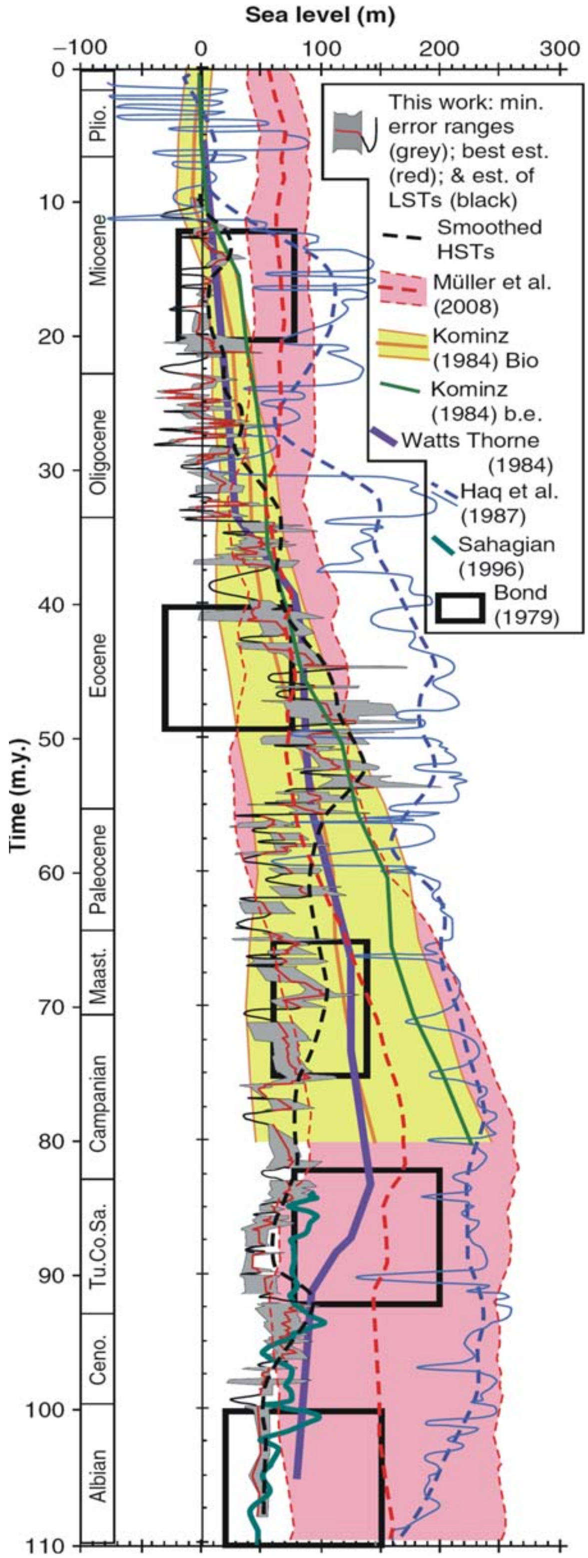
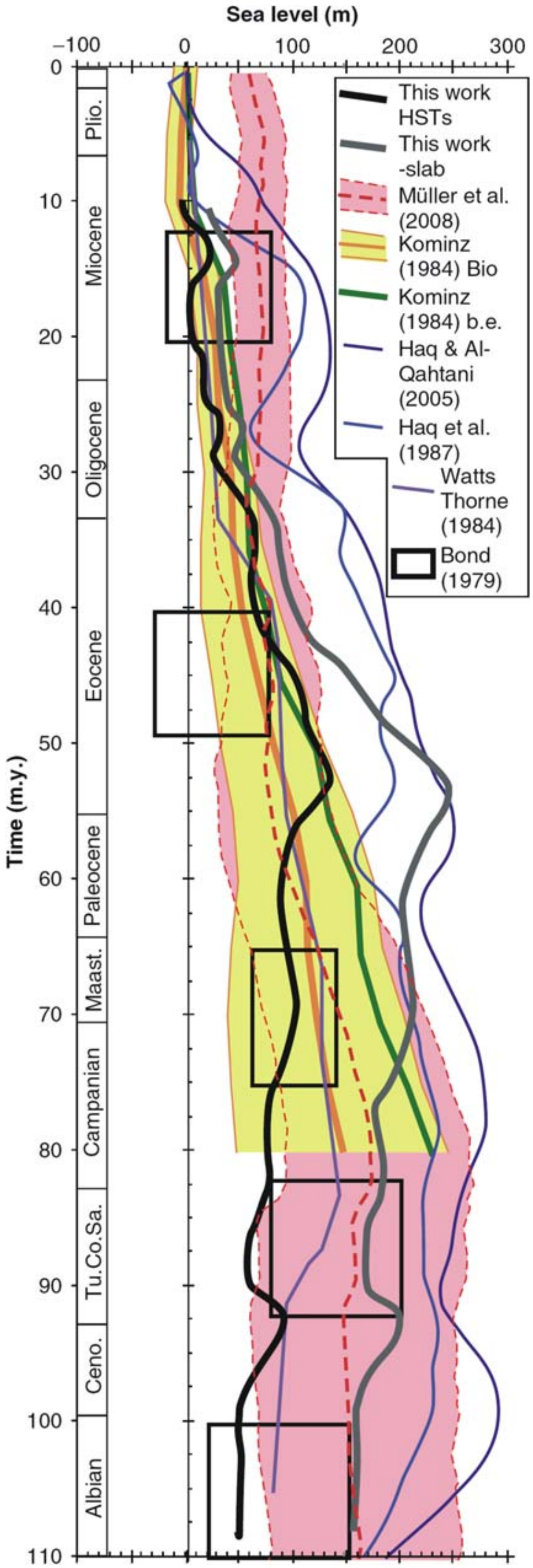


Fig. 9.7 Comparisons of various estimates of eustatic sea-level change. Plio., Pliocene; Ceno., Cenomanian; Tu.Co.Sa, combined Turonian, Coniacian and Santonian; Maast., Maastrichtian (Kominz et al., 2008, Fig. 12)



80 Ma (Kominz, 1984, b.e. curve). The Kominz et al. (2008) curve ("This work" in Fig. 9.7) is based on backstripping analysis of the subsurface stratigraphy of New Jersey, a study examined at some length in Sect. 14.6.1

Kominz et al. (2008) developed a smoothed version of their sea level curve by connecting highstands, and this is shown in Fig. 9.8, against other smoothed, long-term curves. At 80 Ma the discrepancy between the Müller et al. (2008) curve and that of Miller et al. (2005a) against which it was compared, implies a difference in sea level of 130 m. Müller et al. (2008) suggested that during the Late Cretaceous the dynamic load attributed to the subduction of the Farallon Plate (a component of the Panthalassa oceanic crust) beneath North America would have led to regional subsidence of the Atlantic margin of North America. Sea-level estimates based on stratigraphic studies in New Jersey would, therefore, underestimate global sea levels by that amount. Kominz et al. (2008) developed a correction to their new Jersey curve, which is shown in Fig. 9.8 as the curve labelled "This work—slab". The differences between the two Kominz et al. (2008) curves in Fig. 9.8 decrease from Eocene time on, as the dynamic topography effect of the downgoing Farallon Plate is thought to have decreased.

9.3.2 Dynamic Topography and Epeirogeny

Johnson (1971) was one of the first to emphasize the links between Sloss-type cycles of transgression-regression and regional orogeny. Burgess (2008) provided an updated evaluation of this relationship (Fig. 9.9). Johnson (1971) recognized that

> of the four major orogenies that occurred in North America during the Paleozoic and Mesozoic, three began, reached climactic stages, and went through waning stages during the time epicontinental seas were transgressing to their maximum extent and then regressing to form the great onlap-offlap cycles called sequences. . . . The correspondence of orogenic events with onlap of the craton

Fig. 9.8 Comparisons of curves of long-term sea-level change. Ceno., Cenomanian; Tu.Co.Sa, combined Turonian, Coniacian and Santonian; Maast., Maastrichtian; Plio., Pliocene (Kominz et al., 2008, Fig. 13)

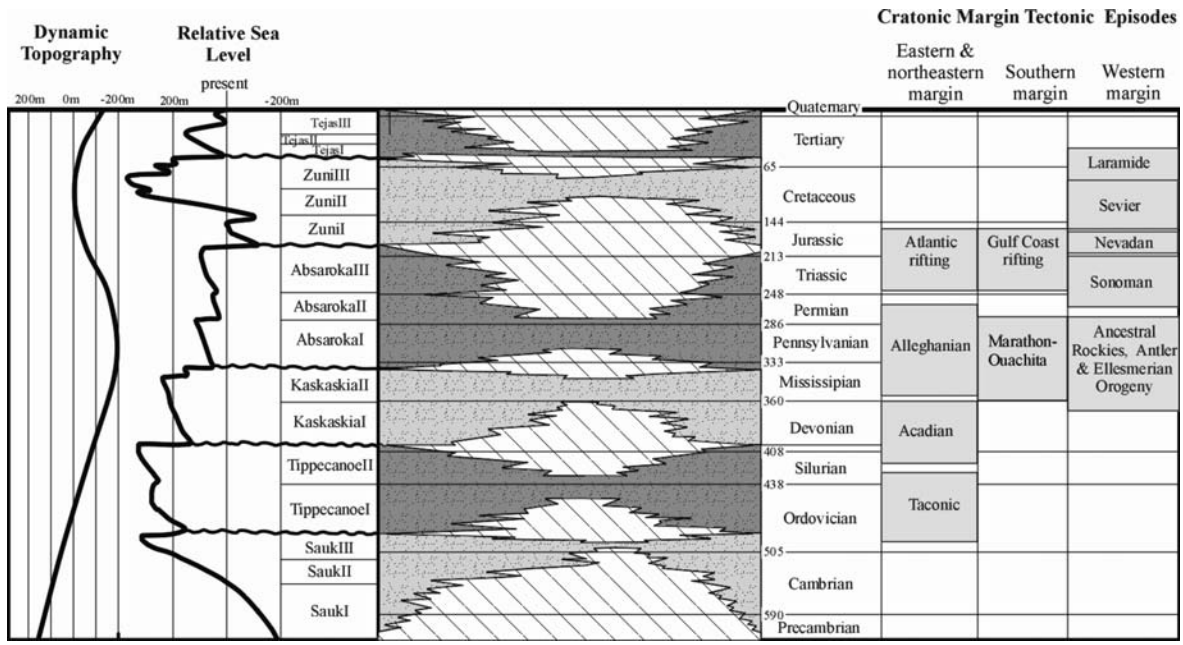


Fig. 9.9 Correlation of the Sloss sequences with craton-margin tectonic episodes (Burgess, 2008)

> is so consistent in general and even in detail, that it must reflect a fundamental relation.

In a series of papers Sloss (1972, 1979, 1982, 1984, 1988b; Sloss and Speed, 1974) elaborated the "Indian-name" sequences that he first established in 1963. He developed ever-more detailed isopach maps and discussed the effects of eustasy and tectonism in the development of the sequences. In one of the earlier of these papers (Sloss and Speed, 1974; see also Sloss, 1984) he attempted to subdivided the six sequences into two broad types. The first type, termed "submergent", and exemplified by the Sauk, Tippecanoe and Kaskaskia sequences,

> . . . is dominated by flexure of the cratonic and interior margins. These sequences exhibit slow regional transgression and onlap, . . . gentle subsidence of interior basins separated by less subsident domes and arches, and a lack of widespread brittle deformation manifested by faulting. The emergent episodes preceding each of these flexure-dominated times of deposition are the occasions for developing of the sequence-bounding unconformities and slow progressive transgression of the craton. (Sloss, 1984, p. 5)

Figure 5.15 illustrates the typical anatomy of the Sauk sequence. The second type of sequence Sloss termed "oscillatory", and

> . . . is characterized by abrupt termination of a submergent episode through rapid cratonic uplift accompanied by high-angle faulting of basement crystalline rocks and, commonly, by faults that propagate from the basement to fracture and displace the overlying sedimentary cover. The Absaroka Sequence represents a typical oscillatory episode. (Sloss, 1984, p. 5)

Examples of these two types of sequence are illustrated by the contrasting subsidence maps of Figs. 5.9 and 5.10. The importance of broad, regional tilting of the craton in the formation of these sequences is apparent from regional cross-sections, such as that shown in Fig. 5.8.

In his later papers, Sloss considered epeirogeny and in-plane intraplate stresses and their effects on sequence architecture. For example, the late Mississippian sub-Absaroka unconformity is a time of pronounced change in cratonic subsidence and uplift patterns within the North American interior. The Canadian Shield emerged as an important source of sediment, and the southwestern Midcontinent (Texas, Oklahoma, New Mexico, Colorado) underwent pronounced submergence. Sloss (1988b) attributed this major change in continental configuration to intraplate stresses associated with plate convergence along the southern margin of the continent. There is now much additional support for these ideas, which are discussed at length here and in the next chapter.

Stratigraphers specializing in the study of continental interiors (e.g., Sloss and Speed, 1974) have for a long time appealed to a process that was termed *epeirogeny* by Gilbert (1890). The modern definition of epeirogeny (Bates and Jackson, 1987) defines it as "a form of diastrophism that has produced the larger features of the continents and oceans, for example plateaux and basins, in contrast to the more localized process of orogeny, which has produced mountain chains." The definition goes on to emphasize vertical motions of the earth's crust.

Modern studies of the thermal evolution of the mantle, supported by numerical modeling experiments, have provided a mechanism that explains the long term uplift, subsidence and tilting of continental areas, especially large cratonic interiors beyond the reach of the flexural effects of plate-margin extension or loading (Gurnis, 1988, 1990, 1992; Burgess and Gurnis, 1995; Burgess, 2008). These studies have shown that the earth's surface is maintained in the condition known as dynamic topography, reflecting the expansion and contraction of the crust resulting from thermal changes in the underlying lithosphere and mantle (Fig. 9.1). Much work remains to be done to test and apply these ideas by developing detailed numerical models of specific basinal stratigraphic histories, although it is already clear that dynamic topography is affected by both upwelling and downwelling currents on several scales. The following paragraphs describe a range of recent studies.

Bond (1978) provided some of the first important insights into epeirogenic processes by demonstrating that the earth's continents have had different histories of uplift and subsidence since Cretaceous time (Sect. 3.4). It is now possible to explain these differences using the concepts of dynamic topography. One of the most striking anomalies revealed by the hypsometric work is the elevation of Australia. "The interior of Australia became flooded by nearly 50% between 125 and 115 Ma and then became progressively exposed between 100 and 70 Ma at a time when nearly all other continents reached their maximum Cretaceous flooding" (Gurnis, 1992). Applying backstripping techniques to detailed isopach maps Russell and Gurnis (1994) estimated that although global sea level was about 180 m above the present level near the end of the Cretaceous, a smaller fraction of the Australian continent was flooded than at the present day. Raising the continent an average of 235 m accounts for the end-Cretaceous paleogeography superimposed on a 180-m-high sea level. This result is related to cessation of subduction on the northeast margin of Australia at about 95 Ma. Subduction had generated a dynamic load by the presence of a cold crustal slab at depth, and the cessation of subduction allowed uplift. Then the northward migration of the continent as it split from Antarctica moved Australia off a dynamic topographic high and geoid low, toward a lower dynamic topography, resulting on continental lowering. Russell and Gurnis (1994) also discussed the broad tilting and basin development of the Australian interior. Burgess and Gurnis (1995) developed preliminary models for Sloss-type cratonic sequences invoking combinations of eustatic sea-level change and dynamic topography.

The thermal consequences of secondary mantle convection above a subducting slab have been invoked as a cause of enhanced subsidence in retroarc foreland basins. Cross (1986) showed how the changing pattern of regional isopachs in the Western Interior foreland basin of the Rocky Mountain region could be explained as a product of the generation of accommodation by crustal flexure caused by the imposition of the supracrustal load of the fold-thrust belt, or a result of regional subsidence over a cool, dense mantle. Mitrovica et al. (1989) developed this idea as an explanation for the anomalously broad extent of the Western Canada Sedimentary Basin during the Cretaceous, and a similar idea has been proposed for the Mesozoic basins of eastern Australia (Gallagher et al., 1994). Although this is clearly a plate-margin effect, the same concepts of dynamic topography apply as in the case of the broader epeirogenic processes discussed above.

> Heat energy is mostly lost from the mantle via convection and consequent volcanism at oceanic spreading centers. Formation of supercontinents that persisted for tens of millions of years, covering a wide region of mantle, prevent this method of heat loss, trapping mantle heat produced by radioactive decay in the core. Such mantle insulation may produce a rise in temperature of ~20 K throughout the mantle (Gurnis and Torsvik, 1994). Thermal expansion due to this temperature increase produces stress on the base of the lithosphere, creating dynamic topography (Anderson, 1982) with an amplitude of ~150 m [Fig. 9.1]. As a consequence of this mechanism, continents should experience uplift during supercontinent formation and persistence, followed by subsidence as the supercontinent breaks up and the fragments drift off hot mantle onto adjacent, relatively cool

> mantle (Anderson, 1982; Gurnis, 1988). Similar effects can be produced by large descending plumes interacting with internal mantle viscosity boundaries (Pysklywec and Mitrovica, 1998) (quote from Burgess, 2008, p. 40).

The major change in cratonic subsidence patterns that occurred during the late Paleozoic (pre-Absaroka), as noted above, occurred at the time the North American continent was impacted by orogenic episodes on three of its four sides, the Alleghanian in the east, the Ouachita-Marathon in the south, and the Ellesmerian in the north (Fig. 9.5). These were three of the important orogenies that culminated in the construction of Pangea.

Burgess (2008, p. 40) pointed out that the two longest-duration lacunae in North America are the base-Sauk and the base-Zuni unconformities. Both formed during periods when North America was part of a supercontinent, Rodinia in the Late Precambrian, and Pangea in the late Paleozoic and early Mesozoic. Cratonic erosion and non-deposition would have been accentuated by dynamic topographic highs creating an emergent craton (Sloss and Speed, 1974). Conversely, the three Paleozoic unconformities are of shorter duration, and formed during a period when North America was one of several dispersed continents, overlying cool mantle, and thus having relatively low or "submergent" (Sloss and Speed, 1974) elevation. Subsidence analysis identifies an anomalously large subsidence event in Late Devonian to Mississippian time, explained by Kominz and Bond (1991) as due to the final stages of assembly of Pangea over a dynamic topographic low (Fig. 9.3).

The mechanism of subduction-induced mantle flow leading to widespread subsidence of foreland basins has been applied to the Carboniferous-Triassic subsidence of the Karoo Basin in southern Africa (Pysklywec and Mitrovica, 1998) and to the Silurian foreland basin adjacent to the Caledonian orogen, now underlying the Baltic Sea (Daradich et al., 2002).

Another importance aspect of dynamic topography is the large-scale continental tilts that can result from plate-margin collisions. Burgess (2008, p. 40) pointed out the absence of Mesozoic cratonic strata in eastern North America, compared with extensive Jurassic and Cretaceous deposition in the west. This may be in part due to slab-related dynamic topography and in part to continent-scale tilting up-to-the-east. Tilting would have resulted from increasing amounts of uplift eastwards across North America towards the hottest mantle situated beneath the center of Pangea.

On the basis of coal moisture measurements and other measures of organic metamorphism, it is known that massive unroofing of the Appalachian and Ouachitan orogens occurred during the late Paleozoic-early Mesozoic. Removal of 4 km of sediment occurred within the undeformed Appalachian foreland basin; there was 7–13 km of cumulative erosion in the central Appalachians, and more than 13 km in the Ouachitas (Beaumont et al., 1988). Bally (1989, p. 430) said: "These amounts of erosion are stunning, and raise the question of the fate of the eroded material and how much of the unloading was due to as yet unrecognised tectonic unroofing." Dickinson (1988) had suggested that much of the thick accumulations of upper Paleozoic and Mesozoic fluvial and eolian strata in the southwestern United States had been derived from Appalachian sources, and this was confirmed by the zircon studies of Dickinson and Gehrels (2003). Neither the location nor scale of specific river systems could be identified by this research, but given the distance of transport and the volume of sediment displaying these provenance characteristics that is now located on the western continental margin, it seems highly probable that large-scale river systems were involved. A westerly tilt of the craton during the late Paleozoic and early Mesozoic could, in part, be attributed to continental-scale thermal doming preceding the rifting of Pangea, followed by the more localized heating and uplift of the rift shoulder of the newly formed Atlantic rift system. Ettensohn (2008) suggested that some of the detritus was also shed eastward, resulting in the accumulation of more than 5 km of sediment in southern New England in Early Jurassic time. The major uplift and erosional unroofing took place during the Late Permian to Early Triassic, and by Late Triassic time "parts of the old orogen in New England had been exhumed to nearly the present erosional level" (Ettensohn, 2008, p. 155; citing isotopic studies of Dallmeyer, 1989).

The extended duration of the sub-Tejas unconformity in eastern North America, and the current elevated topography of North America and the resultant predominantly erosional regime suggest that the dynamic topographic high is somehow persistent or that elevation is being maintained by some other mechanism.

The concepts of dynamic topography have also been invoked to explain cratonic basin formation and subsidence. As noted by Hartley and Allen (1994), there have been at least two major periods in earth history when suites of interior basins formed within large cratons. Both periods are associated with the breakup of supercontinents. The first of these was the early Paleozoic, when the Williston, Hudson Bay, Illinois, Michigan and other basins formed in the cratonic interior of North America. The second period was the Mesozoic breakup of Pangea, when a series of similar basins formed within the continent of Africa. Some of the African basins are undoubtedly related to plate-margin processes, and there has been much debate regarding the importance of reactivation of inherited crustal weaknesses as a cause of the North American basins (Quinlan, 1987). However, Hartley and Allen (1994) suggested that small-scale convective downwelling, decoupled from the large-scale motion, may be a significant factor in basin formation. They found strong evidence for this process in the formation of the Congo Basin. The stratigraphic histories of these suites of cratonic basins are similar but not identical (Quinlan, 1987), and as with the other examples discussed in this section, the processes that maintain dynamic topography are not thought to generate globally simultaneous (eustatic) changes in sea level.

Other examples of the large-scale flux of sediment across the interior of the North American continent, and their possible relationship to dynamic topography, were discussed by Miall (2008).

9.3.3 The Origin of Sloss Sequences

To conclude the discussion in this section: modern ideas about the 10^7-year sequences (the "Sloss sequences": Fig. 5.5) are that they were generated by a combination of two main processes, eustasy and epeirogeny. Cycles of eustatic sea-level change are caused by variations in sea-floor spreading rate superimposed on the supercontinent cycle. These variations are a reflection of continental-scale adjustments to spreading patterns in response to plate rifting and collision events. They occur over time scales in the tens-of-millions of years range, and this can be confirmed by the detailed documentation of spreading histories from the magnetic striping of existing ocean floors, which reveal changes in spreading rate and trajectory over time scales of this magnitude.

Superimposed on the eustatic cycle are continental-scale episodes of warping and tilting driven by mantle thermal processes—the so-called dynamic topography process (Fig. 9.1). Complicating and overlaying both these processes are the regional mechanisms of crustal extension and crustal loading (Sect. 10.2) caused by plate-margin tectonism. In many places, either because the global processes are dominant, or because they are synchronous with regional tectonic episodicity, Sloss's sequences can easily be recognized within the overall stratigraphy. For example, a regional section through western Canada (Fig. 5.4) shows a clear subdivision into Sloss sequences and subsequences (capital letters up the left side of the diagram). Elsewhere, as in the Canadian Arctic Islands, the regional unconformities do not coincide with those defined by Sloss (Fig. 5.7), presumably because they were over-printed by the effects of regional tectonic episodes.

9.4 Main Conclusions

1. A long-term cycle of sea-level change (10^8-year time scale), termed the supercontinent cycle, results from the assembly and breakup of supercontinents on the earth's surface. Eustatic sea-level changes are driven by global changes in sea-floor spreading rate, variations in the average age of the oceanic crust, and variations in continental volumes caused by plate extension and collision.
2. Eustatic sea-level changes on a time scale of tens of millions of years are caused by variations in ocean-basin volume generated by episodic spreading, and by the variations in total length and age of the sea-floor spreading centres as supercontinents assemble, disassemble and disperse.
3. Many other processes have smaller effects on global sea levels through their effects on the volume of the ocean basins. These include oceanic volcanism, sedimentation, ocean temperature changes, and the desiccation of small ocean basins.
4. Epeirogenic effects are "dynamic topography" resulting from thermal effects of mantle convection associated with the supercontinent cycle. This

cycle has long-term consequences, including the generation of persistent geoid anomalies. Vertical continental movements are related to the thermal properties of large- and small-scale convection cells, and can involve continent-wide uplift, subsidence, and tilts, and cratonic basin formation. These movements do not correlate in sign or magnitude from continent to continent.

5. Comparisons of various curves of long-term sea-level change constructed using modern data sets reveal few similarities between curves. It is clear that sea-levels were in the range of at least 200 m higher during the Late Cretaceous than at present, and have slowly fallen since then, but the details of sea-level history even on a $10^{6–7}$-year time scale, remain unresolved.